

Cloud-Scale Numerical Modeling of the Arctic Boundary Layer

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FIRE-III ACCOMPLISHMENT REPORT

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TITLE: Cloud-Scale Numerical Modeling of the Arctic Boundary Layer

ABSTRACT:

The interactions between sea ice, open ocean, atmospheric radiation, and clouds over the Arctic Ocean exert a strong influence on global climate. Uncertainties in the formulation of interactive air-sea-ice processes in global climate models (GCMs) result in large differences between the Arctic, and global, climates simulated by different models. Arctic stratus clouds are not well-simulated by GCMs, yet exert a strong influence on the surface energy budget of the Arctic. Leads (channels of open water in sea ice) have significant impacts on the large-scale budgets during the Arctic winter, when they contribute about 50 percent of the surface fluxes over the Arctic Ocean, but cover only 1 to 2 percent of its area. Convective plumes generated by wide leads may penetrate the surface inversion and produce condensate that spreads up to 250 km downwind of the lead, and may significantly affect the longwave radiative fluxes at the surface and thereby the sea ice thickness. The effects of leads and boundary layer clouds must be accurately represented in climate models to allow possible feedbacks between them and the sea ice thickness.

The FIRE III Arctic boundary layer clouds field program, in conjunction with the SHEBA ice camp and the ARM North Slope of Alaska and Adjacent Arctic Ocean site, will offer an unprecedented opportunity to greatly improve our ability to parameterize the important effects of leads and boundary layer clouds in GCMs.

We will make extensive use of FIRE III measurements and a high-resolution numerical model, the University of Utah Cloud Resolving Model, to increase our understanding of the physical processes that determine:

- (1) the formation and structure of Arctic stratus clouds and
- (2) how leads over the Arctic Ocean affect the large-scale budgets of sensible heat, water vapor, and condensate.

This is necessary before developing lead and boundary layer cloud parameterizations based on general physical principles. We will:

- (1) Determine in detail how large-scale processes, in combination with cloud-scale radiative, microphysical, and dynamical processes, govern the formation and multi-layered structure of Arctic stratus clouds.
- (2) Quantitatively determine the effects of leads on the large-scale budgets of sensible heat, water vapor, and condensate in a variety of Arctic winter conditions.

GOALS: Our long-term research goals are to improve our ability to parameterize the important effects of Arctic leads and Arctic boundary layer clouds in GCMs.

OBJECTIVES: Our FIRE-III research objectives are:

- (1) To determine in detail how large-scale processes, in combination with cloud-scale radiative, microphysical, and dynamical processes, govern the formation and multi-layered structure of arctic stratus clouds. This information will be useful for developing and improving 1D boundary layer models for the Arctic.

- (2) To quantitatively determine the effects of leads on the large-scale budgets of sensible heat,

water vapor, and condensate in a variety of Arctic winter conditions. This information will be used to identify the most important lead-flux processes that require parameterization in climate models.

APPROACH: We made extensive use of a high-resolution numerical model, the 2D University of Utah Cloud Resolving Model (UU CRM), and its 1D version, the University of Utah Turbulence Closure Model (UU TCM), a boundary layer model based on third-moment turbulence closure.

TASKS COMPLETED:

- (1) Our first step toward parameterizing the effects of Arctic leads was to study the effects of cool and dry downdrafts associated with precipitating cumulus clouds on the area-averaged surface turbulent fluxes over a Tropical ocean. As over the Arctic sea ice, the local surface turbulent fluxes may vary significantly over short times and distances.
- (2) We constructed a new 3D large-eddy simulation model which will be used for studying the 3D structure of plumes generated by Arctic leads.
- (3) We ran and analyzed 2D UU CRM simulations of the 3D Glendening and Burk (1992) and 2D Alam and Curry (1994) lead-generated plume cases to gauge the impacts of model physics and dimensionality.
- (4) We used the 2D UU CRM to study how the height of plumes generated by leads in the wintertime Arctic depends on lead width, lead surface buoyancy flux, and lead orientation relative to the large-scale wind direction. We also used the 2D UU CRM to simulate the structure and impact of clouds produced by lead-generated plumes.
- (5) We ran and analyzed 1D UU TCM simulations of the McInnes and Curry (1994) Arctic stratus cloud case to investigate formation mechanisms.
- (6) We also simulated an idealized dry but radiatively driven boundary layer (one filled with radiatively active smoke) with the 1D UU TCM in order to evaluate the model by comparing the results to those from several 1D, 2D, and 3D models that participated in an intercomparison exercise.
- (7) I participated in the field phase of the FIRE-III Arctic Clouds Experiment by flying on two research flights.
- (8) We made improvements to the radiative transfer model used in the UU CRM and UU TCM:
 - (a) In order to model the effects of Arctic wintertime ice clouds on the radiative energy budget, the radiative properties of ice crystals in the infrared are needed. We developed a composite scheme that is valid for nonspherical ice particles over a wide range of size parameters.
 - (b) Because clouds alter the distribution of the solar flux between direct and diffuse components and the penetration of shortwave radiation into the sea ice depends on the ratio of diffuse to direct radiation, we have separated the direct and diffuse components of the solar flux in the radiative transfer model.
- (9) We also developed a broadband 3D radiative transfer model to study the short-wave radiative effects of the horizontally inhomogeneous Arctic stratocumulus clouds over the horizontally inhomogeneous, highly-reflective snow/ice surface.

FUTURE PLANS: We will:

- (1) Use a 3D large-eddy simulation model to study the 3D structure of lead-induced plumes.
- (2) Simulate FIRE-III and SHEBA observations of lead-generated plumes using the 2D CRM.
- (3) Develop a parameterization of the large-scale effects of leads that includes cloud production.
- (4) Use the broadband 3D radiative transfer model in combination with the 2D CRM to study the short-wave radiative effects of the horizontally inhomogeneous Arctic stratocumulus clouds over the horizontally inhomogeneous, highly-reflective snow/ice surface and the interaction of radiation and cloud dynamics.
- (5) Run and analyze 1D TCM, 2D CRM, and 3D large-eddy simulations of Arctic stratus clouds

based on FIRE-III and SHEBA measurements to evaluate and improve the models and to investigate proposed Arctic stratus cloud formation mechanisms.

(6) Run and analyze 1D TCM and 2D CRM simulations of Arctic stratus clouds based on climatological forcing and SHEBA measurements to investigate formation mechanisms and seasonal cloudiness transitions. These will include month-long (or longer) 1D TCM simulations.

(7) Coordinate the processing and assembly of integrated datasets for single-column, cloud-resolving, and large-eddy simulation models based on FIRE-III and SHEBA measurements.

(8) Incorporate the Ebert and Curry sea-ice model into the UU TCM, and the Alam and Curry lead-freezing parameterization into the UU CRM.

RESULTS:

SURFACE FLUXES AND CUMULUS CIRCULATIONS

The primary effect of boundary layer cumulus circulations is to increase the area-averaged wind speed over the speed of the area-averaged vector wind. By neglecting this "gustiness" effect, the area-averaged surface turbulent fluxes may be significantly underestimated. We developed a parameterization of the gustiness that relates it to the cumulus activity (Zulauf and Krueger 1997, 1998). The agreement between the gustiness simulated by the UU CRM and that observed gives us more confidence in the UU CRM's ability to predict the area-averaged effects of lead-induced convective plumes, for which few measurements currently exist.

LEAD-GENERATED PLUMES

We verified that the dependence of the height of lead-generated plumes on lead width and lead orientation relative to the wind direction follows classical similarity solutions (for wind-parallel leads) over a wide range of lead widths, from 200 m to 6400 m, and Glendening and Burk's (1992) scaling law (for leads oriented at an angle to the wind) over a wide range of lead orientations (Zulauf and Krueger 1999).

We also found that the 2D UU CRM produces mean plume characteristics that are quite similar to those from a 3D large-eddy simulation model, and that the addition of microphysical and radiative processes has the greatest impact upon the more energetic plume circulations, such as those associated with wide leads.

SMOKE CLOUD INTERCOMPARISON

An intercomparison of radiatively driven entrainment and turbulence in a smoke cloud (Bretherton et al. 1997) investigated an idealized case in which a convective boundary layer filled with radiatively active "smoke" is simulated by several 1D, 2D, and 3D models. We used this case to test the UU TCM. The results using the UU TCM agree well with the 3D large-eddy simulations involved in the intercomparison (Krueger et al. 1999a).

ARCTIC STRATUS CLOUDS

We investigated the sensitivity of the formation and structure of simulated Arctic summertime boundary-layer clouds to large-scale vertical velocity and drizzle using the UU TCM (Krueger et al. 1999b). The baseline case, under conditions of no large-scale vertical motion, contains two layers of clouds: a stable fog layer near the surface and a stratus cloud in an elevated mixed layer. We found that large-scale subsidence delays the formation of the upper cloud layer, and increases the liquid water content of the fog layer. Neglecting drizzle has little impact on the cloud water mixing ratios in the upper cloud layer, but significantly increases them in the fog layer.

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FIRE-III SIGNIFICANT HIGHLIGHTS

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SIGNIFICANT HIGHLIGHT: Parameterization of Mesoscale Enhancement of Large-Scale Surface Fluxes due to Cumulus Circulations

Large-scale models typically diagnose the surface turbulent fluxes of sensible and latent heat using the large-scale (i.e., area-averaged) near-surface temperature and water vapor mixing ratio and the speed of the large-scale wind vector. These fluxes may be called the "vector-mean" surface fluxes. Esbensen and McPhaden (1996) defined "mesoscale enhancement" as the difference between the vector-mean surface fluxes and the actual large-scale surface fluxes. In the absence of mesoscale circulations, there would be no mesoscale enhancement.

Based on TOGA TAO buoy data, Esbensen and McPhaden found that mesoscale enhancement of evaporation can reach 30% of the total evaporation. They also showed that mesoscale enhancement is primarily due to mesoscale wind variability ("gustiness") and is associated with periods of significant precipitation.

Multi-day, large-domain simulations of tropical convection performed with the 2D University of Utah Cloud Resolving Model indicate that the characteristics of mesoscale enhancement as simulated by the model are quite similar to those observed.

In the accompanying figure, the blue lines show the time series of the simulated large-scale surface wind speed (top panel) and latent heat flux (bottom panel). The green lines show the speed of the simulated large-scale wind vector (top) and the latent heat flux calculated using this speed (bottom). Mesoscale enhancement is the difference between the blue and green lines in the bottom panel.

If the large-scale surface wind speed (blue line, top) is used in place of the speed of the large-scale wind vector (green line, top) to calculate the surface fluxes, the resulting fluxes are practically identical to the actual fluxes (blue line, bottom). Observations support this conclusion, as well.

We have developed a parameterization of the large-scale surface wind speed. It augments the speed of the large-scale wind vector with a gustiness speed that is directly linked to a measure of cumulus activity, (the updraft cumulus mass flux at cloud base level, which is available in many global climate models). The red lines in the figure show the large-scale surface wind speed estimated using the parameterization (top) and the latent heat flux calculated using this speed (bottom). The good agreement between the red and blue lines indicates that the parameterization of mesoscale enhancement is successful.

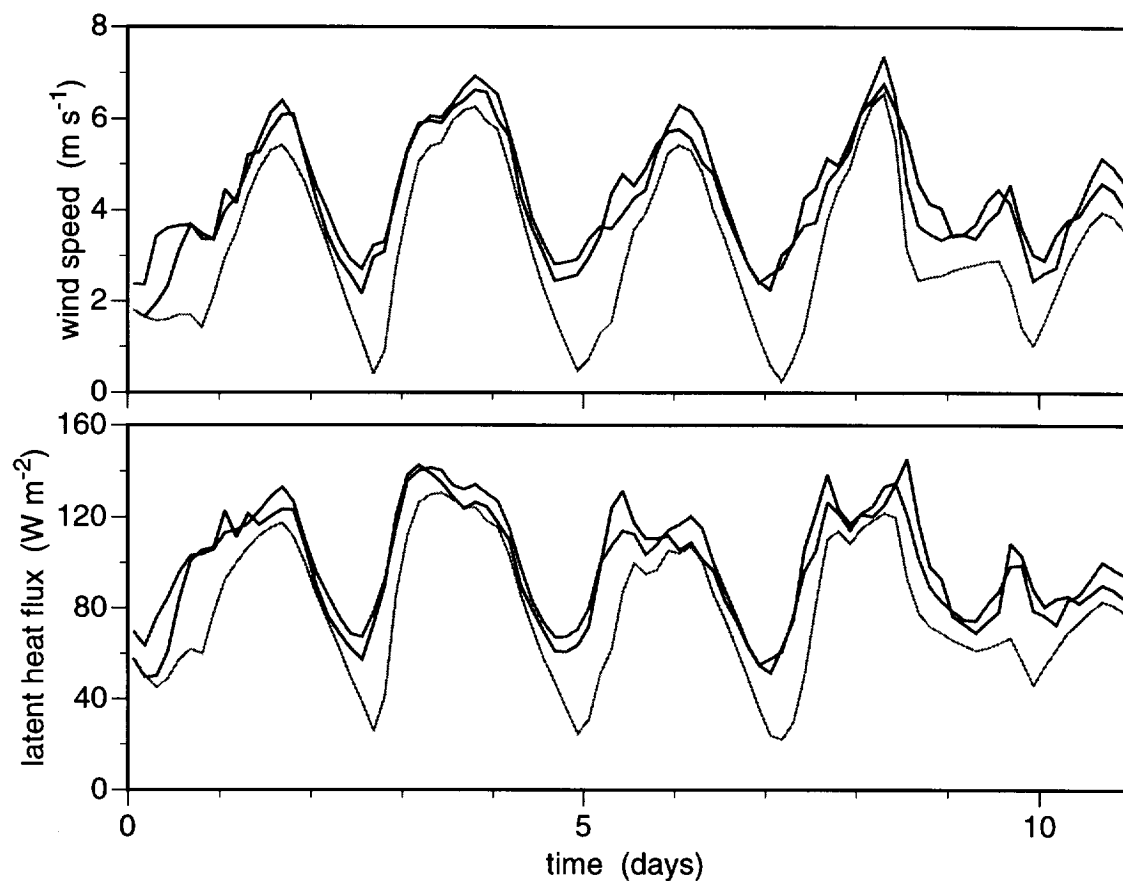


Figure 1: *Blue lines*: simulated large-scale surface wind speed (top) and latent heat flux (bottom). *Green lines*: speed of the large-scale wind vector (top) and the latent heat flux calculated using this speed (bottom). *Red lines*: estimated large-scale surface wind speed based on a parameterization that links gustiness to cumulus activity (top) and the latent heat flux calculated using this speed (bottom).

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SIGNIFICANT HIGHLIGHT: Two-Dimensional Numerical Simulations of Arctic Leads

Leads (channels of open water in sea ice) have significant impacts on the large-scale budgets during the Arctic winter, when they contribute about 50 percent of the surface fluxes over the Arctic Ocean, but cover only 1 to 2 percent of its area. Convective plumes generated by wide leads may penetrate the surface inversion and produce condensate that spreads up to 250 km downwind of the lead, and may significantly affect the longwave radiative fluxes at the surface and thereby the sea ice thickness. The effects of leads must be accurately represented in climate models to allow possible feedbacks between them and the sea ice thickness.

We used a high-resolution numerical model, the University of Utah Cloud Resolving Model (UU CRM), to increase our understanding of the physical processes that determine how leads over the Arctic Ocean affect the large-scale budgets of sensible heat, water vapor, and condensate. This is necessary before developing lead parameterizations based on general physical principles. During FIRE-III:

- We ran and analyzed 2D UU CRM simulations of the 3D Glendening and Burk (1992) and 2D Alam and Curry (1994) lead-generated plume cases to gauge the impacts of model physics and dimensionality.

We found that the 2D UU CRM produces mean plume characteristics that are quite similar to those from a 3D large-eddy simulation model, and that the addition of microphysical and radiative processes has the greatest impact upon the more energetic plume circulations, such as those associated with wide leads.

- We used the 2D UU CRM to study how the height of plumes generated by leads in the wintertime Arctic depends on lead width, lead surface buoyancy flux, and lead orientation relative to the large-scale wind direction.

We verified that the dependence of the height of lead-generated plumes on lead width follows classical similarity solutions (for wind-parallel leads) over a wide range of lead widths, from 200 m to 6400 m, (see top panel of figure), and that the dependence on lead orientation relative to the wind direction follows Glendening and Burk's (1992) scaling law (for leads oriented at an angle to the wind) over a wide range of lead orientations (Zulauf and Krueger 1999) (see bottom panel of figure).

These results not only indicate that the 2D UU CRM will be a useful for determining how leads over the Arctic Ocean affect the large-scale budgets of sensible heat, water vapor, and condensate, but that knowledge of the lead width and orientation distributions are needed in order to parameterize the effects of leads on the atmosphere and on the surface heat budget.

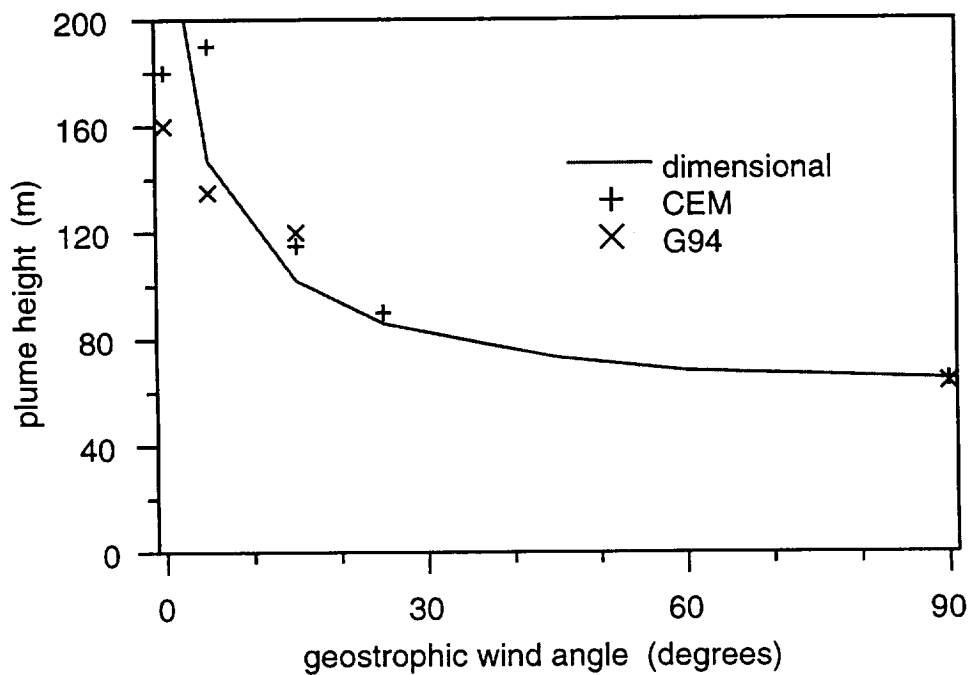
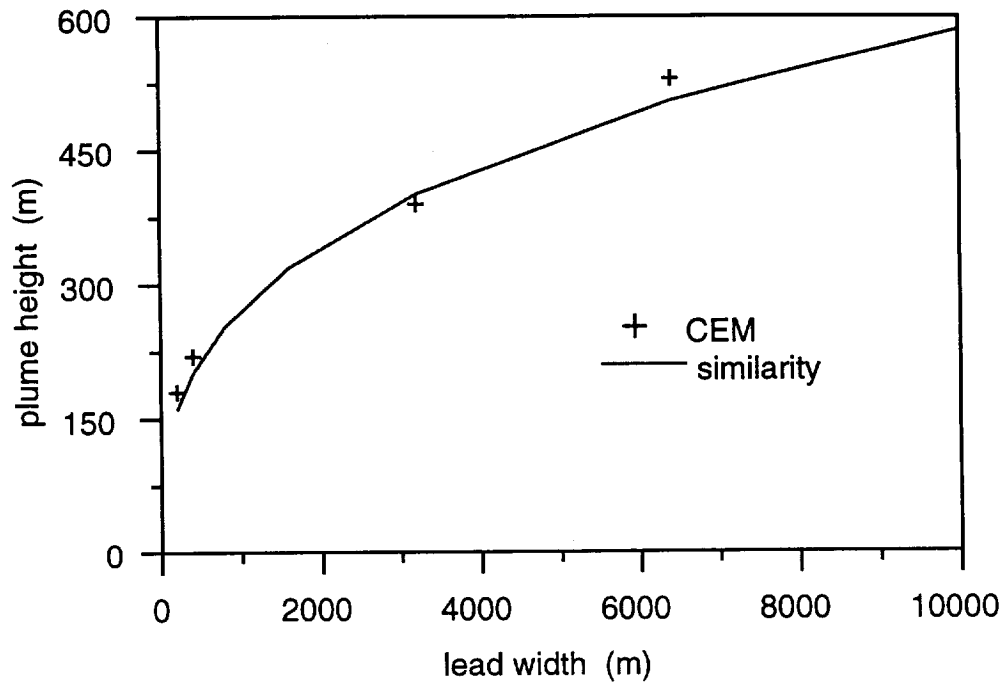


Figure 1: Comparison of simulation derived plume heights with theoretical solutions for cases with specified surface fluxes and no geostrophic cross-wind (top), and those with diagnosed surface fluxes and geostrophic cross-wind (bottom).

FIRE-III SIGNIFICANT HIGHLIGHTS

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SIGNIFICANT HIGHLIGHT: Three-Dimensional Numerical Simulations of Arctic Leads

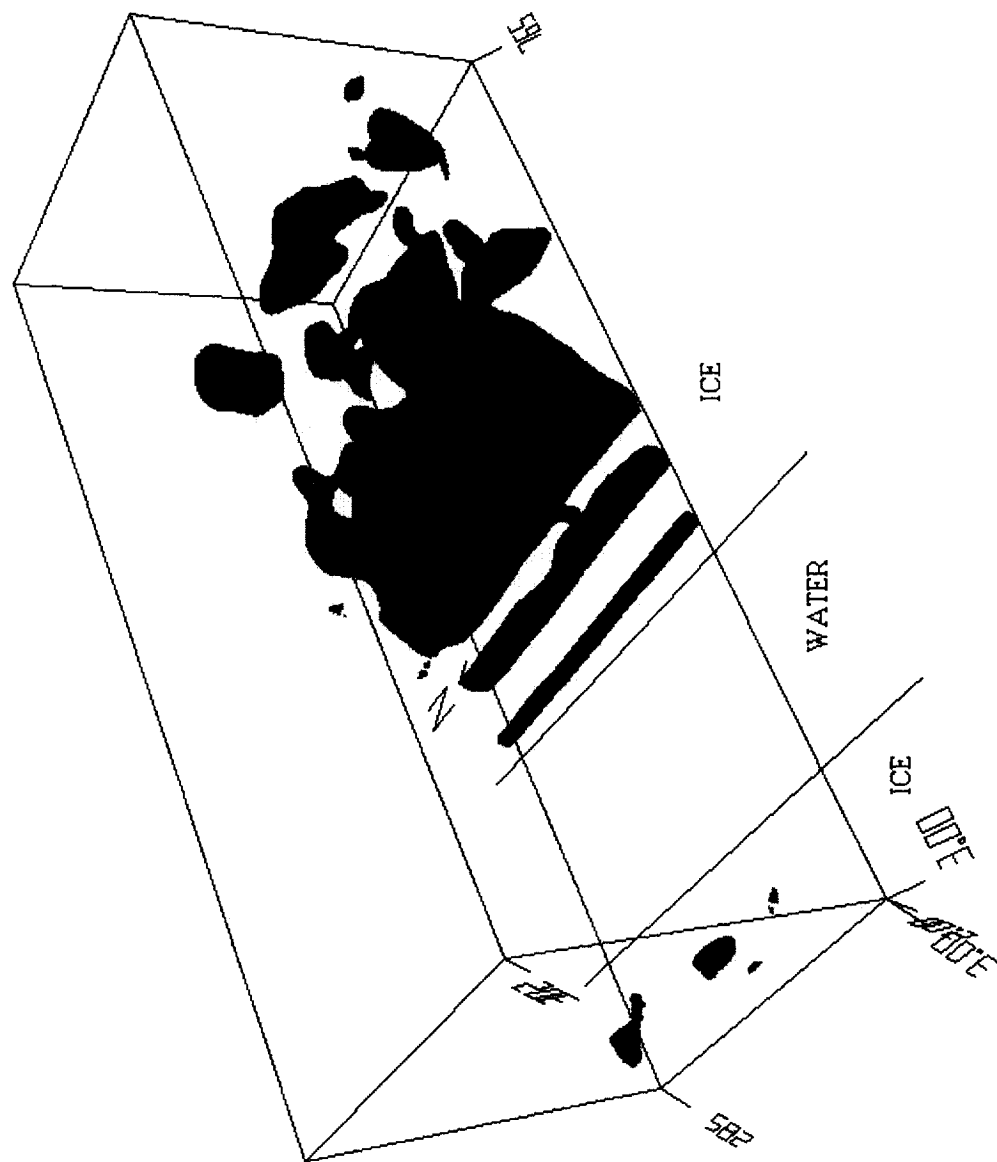
Leads (channels of open water in sea ice) have significant impacts on the large-scale budgets during the Arctic winter, when they contribute about 50 percent of the surface fluxes over the Arctic Ocean, but cover only 1 to 2 percent of its area. Convective plumes generated by wide leads may penetrate the surface inversion and produce condensate that spreads up to 250 km downwind of the lead, and may significantly affect the longwave radiative fluxes at the surface and thereby the sea ice thickness. The effects of leads must be accurately represented in climate models to allow possible feedbacks between them and the sea ice thickness.

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During FIRE-III:

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- We used the 2D UU CRM to study how the height of plumes generated by leads in the wintertime Arctic depends on lead width, lead surface buoyancy flux, and lead orientation relative to the large-scale wind direction. We also used the 2D UU CRM to simulate the structure and impact of clouds produced by lead-generated plumes.
- We constructed a new 3D large-eddy simulation model which will be used for studying the 3D structure of plumes generated by Arctic leads. In particular, we will use it to evaluate the 2D UU CRM simulations.

The figure shows a snapshot from a 3D large-eddy simulation of a lead-generated convective plume using our new model. The 20 cm/s vertical velocity iso-surface is displayed. On the scale of the lead width, the plume's updraft structure is essentially two-dimensional. However, on smaller scales many 3D turbulent eddies are evident. These eddies are responsible for most of the vertical transport of heat and water vapor, so their effects must be accurately represented in the 2D CRM.



Snapshot from a University of Utah 3D large-eddy simulation of a lead-generated convective plume. The 20 cm/s vertical velocity iso-surface is displayed. The air temperature is -27°C , the ice temperature is -29°C , and the water temperatures is -2°C .

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SIGNIFICANT HIGHLIGHT: Simulation of a Summertime Arctic Cloudy Boundary Layer Using a 1D Turbulence Closure Model

Arctic stratus clouds (ASC) significantly affect the surface energy budget of the Arctic, and thereby the sea ice thickness. The formation mechanism(s) of these clouds are still uncertain. ASC often occur in multiple layers. The lowest layer may rest on the surface. The upper cloud layer or layers are decoupled from the stable surface layer, thus their source of water vapor does not appear to be the surface.

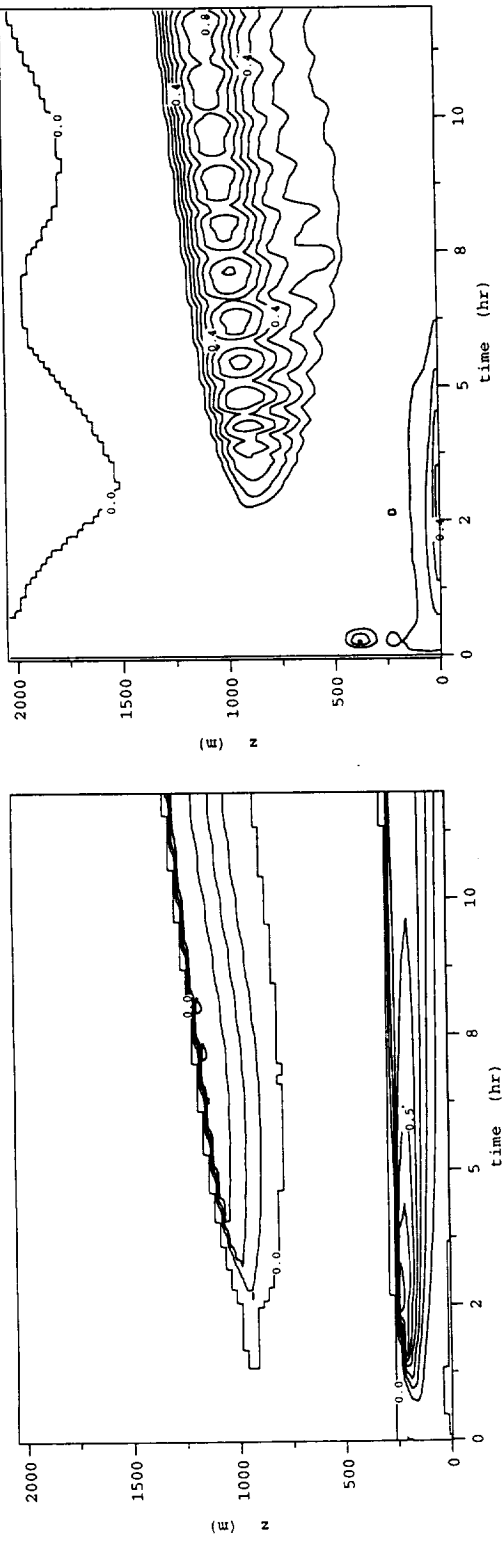
Several layering mechanisms have been proposed:

- Herman and Goody (1976): Solar absorption warms the interior of a surface-based cloud and thereby forms a clear layer between two cloudy layers.
- Tsay and Jayaweera (1984): The upper cloud layer is formed by very weak ascent or entrainment, while the surface cloud layer is an advective fog.
- McInnes and Curry (1995): The upper cloud layer is maintained by cloud-top radiative cooling, while the lower cloud layer is formed by radiative cooling at the base of the upper mixed layer.

1D second-moment turbulence closure models have reproduced many aspects of the structure of the summertime Arctic boundary layer. We used the University of Utah 1D third-moment turbulence closure model (which includes a simple drizzle parameterization and an interactive radiative transfer scheme) to simulate an Arctic summertime boundary layer based on observations obtained on June 28, 1980, during the Arctic Stratus Experiment (McInnes and Curry 1995).

The figure shows time-height plots of cloud water mixing ratio and turbulent kinetic energy from a simulation with no drizzle and no large-scale vertical motion. (Including drizzle makes little difference.) There are two cloud layers: a stable fog layer near the surface and a stratus cloud in an elevated turbulently mixed layer. The upper cloud layer slowly ascends due to entrainment. Both layers form due to cloud-top radiative cooling; the fog layer is also cooled by contact with the surface.

The ability of this model to simulate the formation and structure of a summertime Arctic cloudy boundary layer suggests that it should be useful for studies of boundary layers observed during the FIRE-III Arctic Cloud Experiment.



Time-height plots of cloud water mixing ratio (g kg^{-1}) (left) and turbulent kinetic energy ($\text{m}^2 \text{s}^{-2}$) (right) from a simulation of an Arctic summertime boundary layer using the University of Utah 1D turbulence closure model. The simulation is based on observations obtained on June 28, 1980, during the Arctic Stratus Experiment.

APPENDIX: SELECTED PREPRINTS

- Krueger, S. K., and M. Zulauf, 1997: Parameterization of mesoscale enhancement of large-scale surface fluxes over tropical oceans. *Preprints, 12th Symposium on Boundary Layers and Turbulence*, Vancouver, B.C., Canada, Amer. Meteor. Soc., 494–495.
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1. INTRODUCTION

Large-scale models typically diagnose the surface turbulent fluxes of sensible and latent heat over the ocean using the large-scale (i.e., area-averaged) near-surface temperature and water vapor mixing ratio and the speed of the large-scale wind vector. These fluxes may be called the “vector-mean” surface fluxes. Esbensen and McPhaden (1996) defined “mesoscale enhancement” as the difference between the vector-mean surface fluxes and the actual large-scale surface fluxes. In the absence of mesoscale circulations, there would be no mesoscale enhancement.

Based on TOGA TAO buoy data, Esbensen and McPhaden found that mesoscale enhancement of evaporation can reach 30% of the total evaporation. They also showed that mesoscale enhancement is primarily due to mesoscale wind variability (“gustiness”) and is associated with periods of significant precipitation.

Analyses of two multi-day, large-domain simulations of tropical maritime convection performed with the 2D UCLA cumulus ensemble model (CEM) indicate that the characteristics of mesoscale enhancement as simulated by the model are quite similar to those observed. Xu et al. (1992) describe the two simulations (Q02 and Q03) in detail. Each simulation was for 11 days in a 512-km domain with a horizontal grid size of 2 km.

2. SURFACE FLUX CALCULATIONS

In the CEM, the bulk aerodynamic method of Dardorff (1972) is used to calculate the local (i.e., gridpoint) latent and sensible heat fluxes at the surface, E and S . Using the notation of Esbensen and McPhaden, the domain-averaged latent heat flux, $\langle E \rangle$, for example, is then

$$\langle E \rangle = \rho L_v \langle C_q(U, \cdot) U (q_s - q) \rangle,$$

where ρ is the density, L_v is the latent heat of vaporization, C_q is the stability-dependent transfer coefficient, U is the wind speed, q_s is the surface mixing ratio, q is the near-surface mixing ratio, and angle brackets indicate the domain average. The corresponding vector-mean flux is

$$E_v = \rho L_v C_q(V, \cdot) V (\langle q_{sc} \rangle - \langle q \rangle),$$

where V is the magnitude of the average (or resultant) wind vector, which is predicted by large-scale models.

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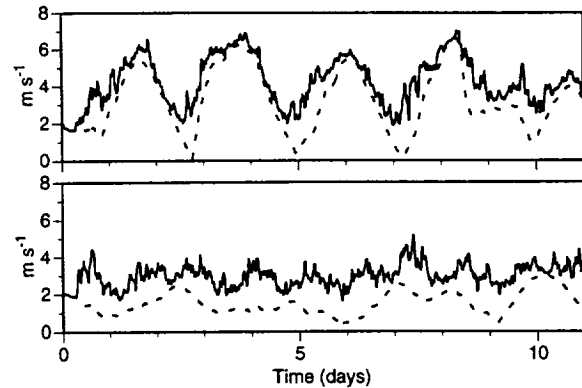


Figure 1. For Q02 (top) and Q03 (bottom): $\langle U \rangle$ (solid) and V (dashed).

By replacing V with the domain-averaged wind speed, $\langle U \rangle$, we obtain the scalar-mean flux,

$$E_s = \rho L_v C_q(\langle U \rangle, \cdot) \langle U \rangle (\langle q_s \rangle - \langle q \rangle).$$

As can be seen in Fig. 1, the differences between $\langle U \rangle$ and V can be significant. Fig. 2 compares the vector-mean and scalar-mean latent heat fluxes with the domain-averaged flux. The results are similar for the corresponding sensible heat fluxes. We see that mesoscale enhancement is significant, and that the vector-mean flux does a poor job of estimating the averaged flux, while the scalar-mean flux tracks it extremely well. Table 1 summarizes these results.

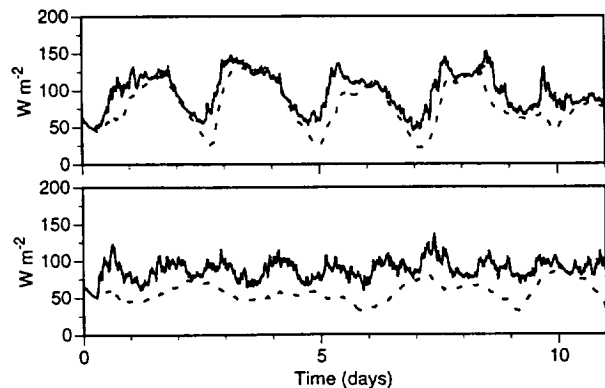


Figure 2. For Q02 (top) and Q03 (bottom): $\langle E \rangle$ (solid), E_v (dashed), and E_s (dotted).

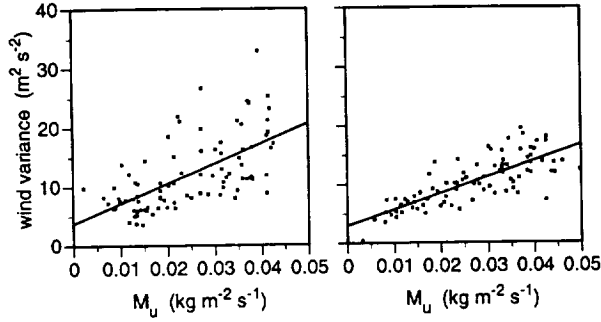


Figure 3. For Q02 (left) and Q03 (right): scatter plots of M_u at $z=566$ m vs. U_σ^2 .

3. WIND SPEED PARAMETERIZATION

We can approximate the scalar-mean fluxes, which are good estimates of the large-scale (i.e., domain-averaged) fluxes, if we can parameterize $\langle U \rangle$ in terms of V . Since V is always less than or equal to $\langle U \rangle$, we may introduce a gustiness speed, U_g , such that

$$\langle U \rangle^2 = V^2 + U_g^2. \quad (1)$$

The variance of the horizontal wind speed,

$$U_\sigma^2 \equiv \langle u'^2 \rangle + \langle v'^2 \rangle,$$

where u' and v' are the local deviations from $\langle u \rangle$ and $\langle v \rangle$, the components of the average wind vector, can be formally related to the gustiness speed:

$$U_g = \alpha U_\sigma. \quad (2)$$

Theoretical methods give a value of 0.8 for α (Jabouille et al. 1996). In our simulations, α is usually between 0.75 and 0.85. These results suggest that a constant value of α is a good approximation. Then U_g , and hence $\langle U \rangle$, may be obtained if U_σ can be parameterized.

We will assume that the wind speed variance is primarily due to mesoscale circulations driven by cumulus convection. If this is the case, U_σ^2 should be related to a measure of the intensity of the cumulus convection, such as the updraft cloud mass flux, M_u . Typical relationships between the two are shown in Fig. 3, where 3-hour averages of M_u at cloud-base level and U_σ^2 from Q02 and Q03 are related through scatterplots. In each instance, a least squares fit to a line was performed. The fit for Q02 has a correlation coefficient of 0.39, while that for Q03 is substantially better, with a correlation coefficient of 0.61. The differences reflect differences in the mesoscale organization of convection between Q02 and Q03.

The linear fits displayed in Fig. 3 form the basis of a parameterization of $\langle U \rangle$ that augments V with a gustiness speed that is linked via U_σ^2 to a measure of cumulus activity, M_u at cloud-base level, which is available in many large-scale models.

Table 1 lists the time-average and rms errors for large-scale fluxes obtained using various parameterization methods. The first line of each part of the table

Table 1. Time-averages and rms errors (W/m^2) for large-scale fluxes from various parameterizations.

Case	Method	rms		rms	
		$\langle E \rangle$	$\langle S \rangle$	$\langle E \rangle$	$\langle S \rangle$
Q02	$\langle E \rangle, \langle S \rangle$	99.5	0.0	17.4	0.0
	E_v, S_v	80.4	24.8	13.7	5.2
	E_s, S_s	100.4	2.0	17.4	0.8
	$U_g \leftarrow U_\sigma$	101.0	3.7	17.5	0.9
	$U_g \leftarrow [U_g]$	98.1	9.7	16.9	2.3
	$U_\sigma^2 \leftarrow M_u$	100.8	7.9	17.4	1.7
Q03	$\langle E \rangle, \langle S \rangle$	88.2	0.0	17.5	0.0
	E_v, S_v	57.4	33.4	11.1	6.9
	E_s, S_s	90.5	2.4	17.7	0.4
	$U_g \leftarrow U_\sigma$	91.0	3.5	17.8	0.6
	$U_g \leftarrow [U_g]$	89.6	8.6	17.5	1.7
	$U_\sigma^2 \leftarrow M_u$	90.7	6.4	17.7	1.2

shows the actual domain-averaged fluxes, time-averaged over the last 10.5 days of the 11-day runs. The second line shows the vector-mean fluxes, and the third line the scalar-mean fluxes. The fourth line shows an approximation of the scalar-mean fluxes obtained using (1) and (2) with a constant value for α (the theoretical value of 0.8) along with the actual U_σ .

The fifth line of each table presents approximate scalar-mean fluxes obtained using (1) and the time average of U_g (2.5 m s^{-1} in both runs). For both Q02 and Q03, this method recovers the mean values of the fluxes rather well, but the lack of any time variation in the gustiness leads to large rms errors. Finally, the sixth line shows the approximate scalar-mean fluxes obtained using (1), (2), and the parameterization for wind speed variance in terms of updraft cloud mass flux shown in Fig. 3. The rms errors are less than those for the average U_g method, while the mean errors are similar to those for the actual scalar-mean fluxes.

ACKNOWLEDGMENTS. This research was supported by NASA Grant NAG 1-1718.

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1. INTRODUCTION

Arctic leads, due to the extreme temperature differences between the sea surface and the winter atmosphere, can be a significant source of moisture and heat for the polar atmosphere. Because of their relatively small scales however, these openings in the pack ice cannot be explicitly resolved by large-scale models.

Previous studies have examined some of the various factors which influence the convective plumes that develop in the vicinity of leads. Glendening and Burk (1992) and Glendening (1994), hereafter referred to as GB92 and G94 respectively, examined the impact of orientation of the large-scale wind field over a 200 m wide lead. GB92 and G94 also evaluated the heat transport budget extensively. They conducted three-dimensional large-eddy simulations (LES), wherein the model had been modified to handle the stable stratification of the Arctic winter atmosphere. Alam and Curry (1995, hereafter AC95) used a two-dimensional model to investigate the impacts of lead width and large scale wind field upon plume evolution.

In the present study, the two-dimensional cloud resolving model (CRM) of Krueger, McLean, and Fu (1995) is employed to attempt to verify, and build upon the previous findings. The sensitivity to ambient wind speed and orientation, and lead width is investigated. Comparisons are made with the previous LES and two-dimensional studies, and with results obtained from similarity and dimensional theory. For the most part, previous studies have not included physical processes such as latent heating, liquid/ice microphysics, and radiation. The CRM can be used to help gauge the relative importance of these processes.

By examining the atmospheric circulations created by the leads, we hope to obtain a better understanding of the impact the enhanced small-scale surface fluxes can have upon the large-scale. This will help facilitate an improved understanding of how Arctic leads influence the Arctic climate.

2. SIMULATION PARAMETERS

The model parameters are essentially the same as those employed by GB92 and G94. The atmospheric

base state is stably stratified with potential temperature increasing at 10 K km^{-1} , from a surface air temperature of -27°C . The ice and water surface temperatures are -29°C and -2°C , respectively. The large-scale geostrophic wind is fixed at 2.5 m s^{-1} , while its direction with respect to the lead varies depending upon the simulation. Surface roughness values of 0.1 and 0.01 cm are used for the ice and water, respectively. The Coriolis parameter is calculated at a latitude of 79°N . Unless otherwise noted, all simulations neglect microphysical and radiative processes.

In the experiments presented herein, the horizontal domains range over more than two orders of magnitude. The vertical domains vary to a lesser extent than the horizontal, and are typically deep enough to ensure that wave reflection off of the model top is not a significant problem. The grid spacing and time step vary depending upon the simulation. The lead-perpendicular boundaries are periodic, and the top boundary allows no penetration. To better quantify turbulent and time varying flow fields, all results are time averaged over a specified period near the end of each simulation. Run times span a range from approximately 12 minutes to 6 hours.

3. EFFECTS OF VARYING WIND ORIENTATION

As stated in G94, plume development depends strongly upon transit time of an air parcel across the lead. When the time period is less than the Brunt Väisälä period P , maximum plume development will occur downstream of the lead. When the transit time is greater than the Brunt Väisälä period, maximum plume development occurs over the lead.

Fig. 1 compares the vertical turbulent temperature flux obtained by GB92 and the present study, for a 200 m wide lead with the geostrophic wind oriented at a 90° angle to the lead. Note that, as is the case with many of the figures presented, only a portion of the computational domain is displayed. For this case the transit time is approximately $P/4$, which is not enough time for a well developed and distinct plume to form over the lead. While the shape of the LES and CRM plumes differ substantially, the magnitude of the fluxes and the depth of penetration show excellent agreement. For both simulations the plume penetrates to a depth of about 65 m.

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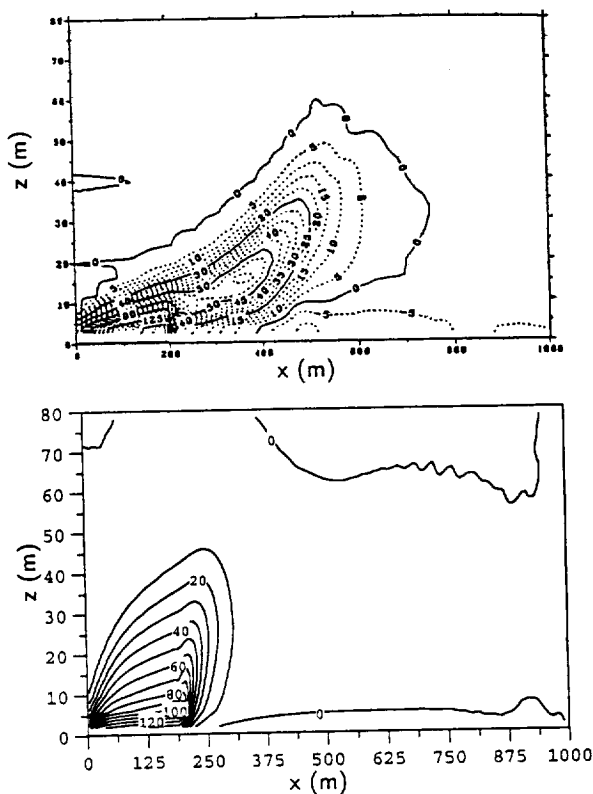


Figure 1: Mean total vertical turbulent temperature flux (K m s^{-1} , scaled by 10^3) for a 200 m lead, 90° wind orientation: GB92 (top), CRM (bottom).

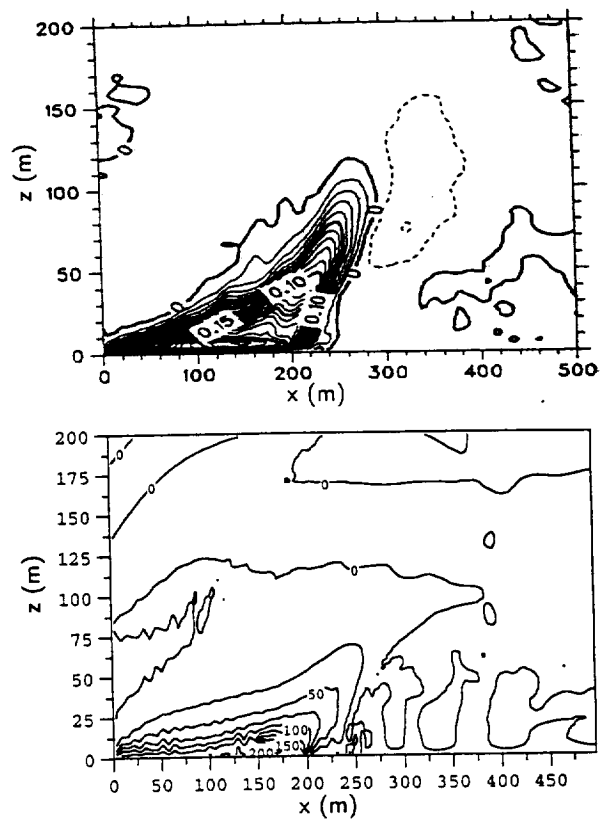


Figure 2: Mean total vertical turbulent temperature flux (K m s^{-1}) for a 200 m lead, 15° wind orientation: G94 (top), CRM (bottom, scaled by 10^3).

Fig. 2 displays the temperature fluxes obtained when the geostrophic wind is aligned at a 15° angle with respect to the lead. Despite the longer transit time, slightly less than P , the maximum plume development remains downstream of the lead, much like the 90° case. The resulting plumes do penetrate deeper (to approximately 120 m). Compared with the CRM, however, the LES displays larger values of the temperature flux reaching slightly higher in the atmosphere, but otherwise the resulting plumes are quite similar.

In Fig. 3 the geostrophic wind is parallel to the lead. Even though there is no large-scale cross-wind, frictional turning of the geostrophic forcing near the surface (the Ekman spiral) causes both plumes to be tilted slightly to the left. It is obvious that the lack of an imposed cross-wind allows the plumes to transport heat much higher into the inversion. The magnitude of the fluxes agree quite well between the two simulations, while the CRM allows the plume to penetrate slightly higher than the LES (approximately 175 m vs. 160 m).

4. EFFECTS OF VARYING LEAD WIDTH

Lead width variation can impact the resulting circulations in a number of ways. The primary effect appears to be that by increasing the warm water surface area, more heat is transferred to the atmosphere, which in turn enhances the local circulations. Among the secondary effects is the fact that wider leads have longer transit times for a given wind field, which facilitates plume development. Another secondary effect is that a wider lead, with its enhanced circulation and inflows, will display greater surface heat fluxes. The increased fluxes further intensify the circulations, forming a positive feedback that can amplify substantially the resulting plumes. The lead fraction (ratio of lead surface area to total surface area) also has an impact upon the circulations. An increase in lead fraction expedites the plume's erosion of the inversion, thereby allowing the plumes to attain greater heights.

In Fig. 4, vertical velocities are displayed for three CRM simulations, for 200 m, 400 m, and 800 m wide leads. Because the geostrophic wind is parallel to the leads (to simplify the analysis) and the plumes develop

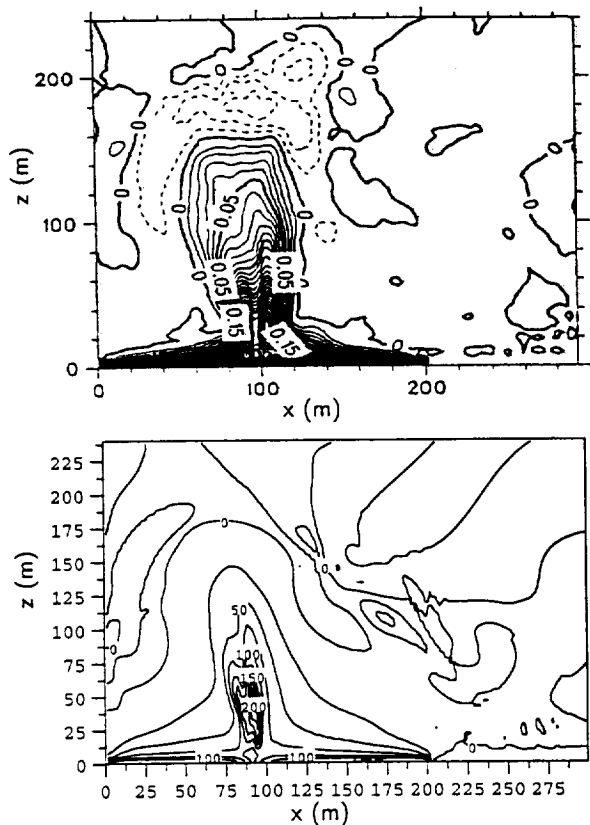


Figure 3: Mean total vertical turbulent temperature flux (K m s^{-1}) for a 200 m lead, 0° wind orientation: G94 (top), CRM (bottom, scaled by 10^3).

over the leads, the transit time is not a factor. The lead fraction is approximately the same in the three plots, (though substantially different from that used in earlier figures). The difference in the plumes displayed in Fig. 4 is thus mainly a result of the increase in warm water surface area, and thereby an increase in the heat transfer to the atmosphere. As expected, the wider leads result in deeper and more vigorous circulations. The plumes attain depths of approximately 175 m, 200 m, and 300 m for the 200 m, 400 m, and 800 m wide leads, respectively. The maximum vertical velocities increase in a similar fashion.

The surface sensible heat fluxes average approximately 240 W m^{-2} over the 200 m lead, 250 W m^{-2} over the 400 m lead, and 270 W m^{-2} over the 800 m lead. Though these increases are modest and probably do not affect plume development greatly, Fig. 5 displays a case where the surface flux feedback is more substantial. For this case, in a quiescent environment, the average sensible heat flux over a 10 km wide lead is 422 W m^{-2} (CRM results). Also shown in Fig. 5 are results from AC95 for a 10 km lead using the same parameters. Note that the plot from AC95 displays their

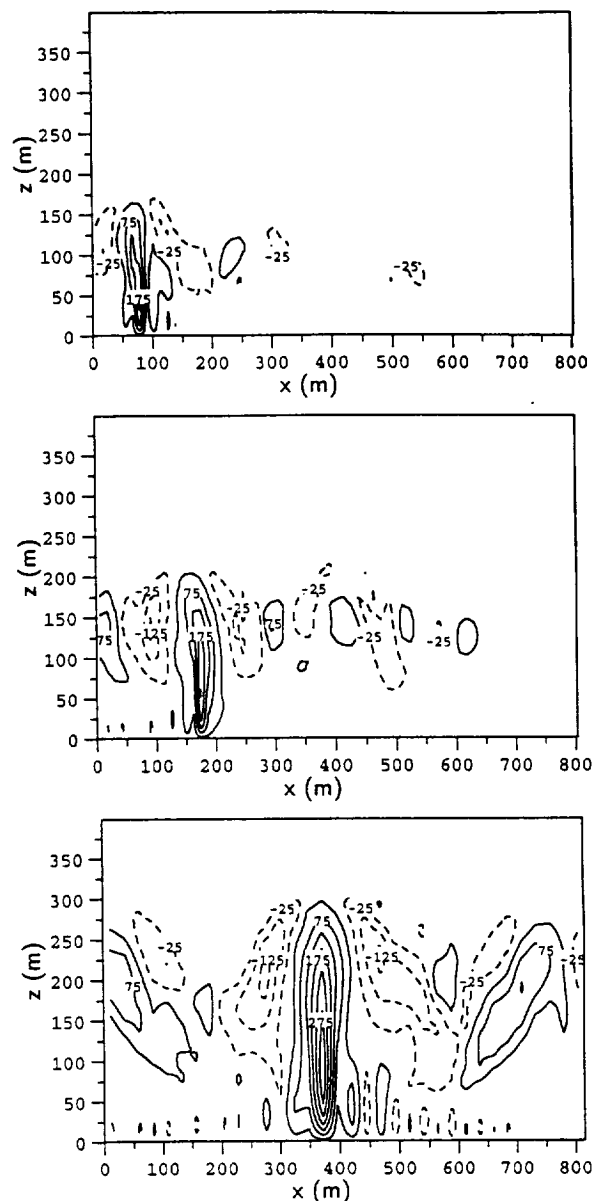


Figure 4: Mean vertical velocity (cm s^{-1}) for 0° wind orientation: 200 m lead (top), 400 m lead (middle), 800 m lead (bottom).

stretched vertical coordinate, and that the units on the axes are in km rather than m.

Fig. 5 compares the vertical velocities obtained after six hours of integration. Note the striking differences between the results of AC95 and the present CRM results, especially plume height. AC95 found that the plume penetrated to an altitude of approximately 2 km, while in the present study the CRM plume height is less than 800 m. Also, after 6 hours, the AC95 plume appeared to still be growing at a substantial rate, while the CRM results indicate slight growth. A possible ex-

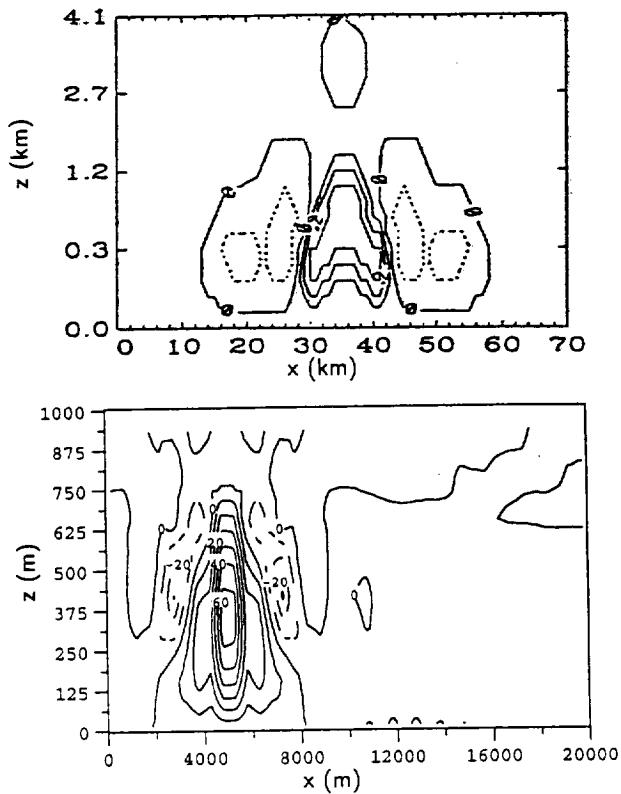


Figure 5: Mean vertical velocity (cm s^{-1} , with 10 cm s^{-1} contour intervals) for a 10 km lead, no geostrophic wind: AC95 (top), CRM (bottom).

planation for the disparities between these results is the more advanced turbulence closure scheme employed by the CRM.

5. COMPARISONS WITH THEORETICAL RESULTS

Additional evidence that the CRM is accurately predicting plume height may be obtained by comparing model results with those from theoretical methods. A similarity solution like those of Emanuel (1994) may be obtained for the case where the large-scale geostrophic wind is parallel to the lead. To eliminate any feedbacks caused by the increased surface flux, the heat flux over the lead is specified at 250 W m^{-2} . The top portion of Fig. 6 compares CRM results with those obtained from similarity theory. Although specifying the fluxes, rather than diagnosing them normally, affects the plume circulations, the excellent agreement with theory leads us to believe that we are handling the basic physics of the problem accurately.

Comparisons with theory become more complicated when the geostrophic cross-wind is non-negligible. Fig. 6 also compares the theoretical heights calculated using the dimensional solution given by GB92 with

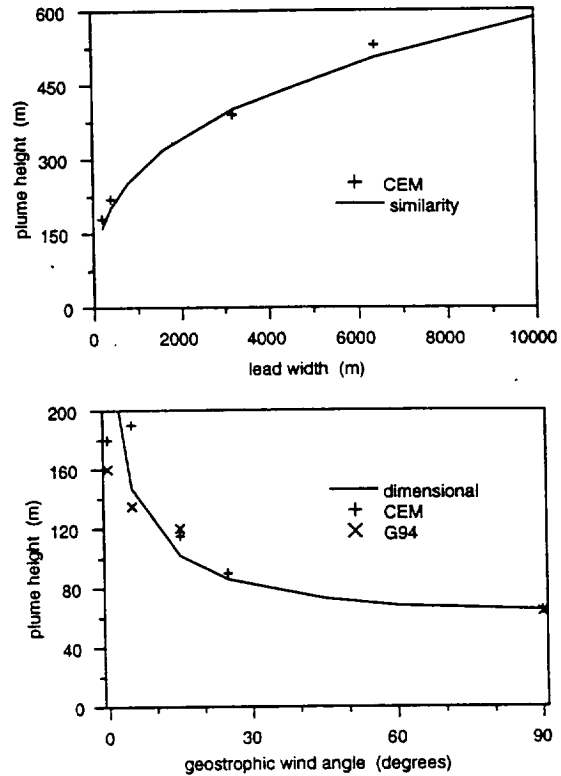


Figure 6: Comparison of simulation derived plume heights with theoretical solutions for cases with specified fluxes and no geostrophic cross-wind (top), and those with natural fluxes and geostrophic cross-wind (bottom).

CRM results for this situation. A 200 m lead with a 2.5 m s^{-1} geostrophic wind at varying wind-to-lead orientation is used. The surface sensible heat fluxes are diagnosed, rather than prescribed, and are all approximately 250 W m^{-2} (the value used for the dimensional solution). For comparison, the LES results from G94 are included on the plot. The excellent agreement between theory, LES, and CRM results suggests that the plumes are being handled adequately.

6. EFFECTS OF ADDITIONAL PHYSICS

For reasons of computational requirements or complexity, many studies have neglected microphysical and radiative processes. For the most part, this is a valid simplification, because at the temperatures and time scales of interest, sensible heating is thought to be the dominant process. Nonetheless, these other processes should also be investigated, especially since the addition of moisture to the Arctic atmosphere is of primary interest. As we are presently interested in the Arctic winter, solar radiation need not be considered.

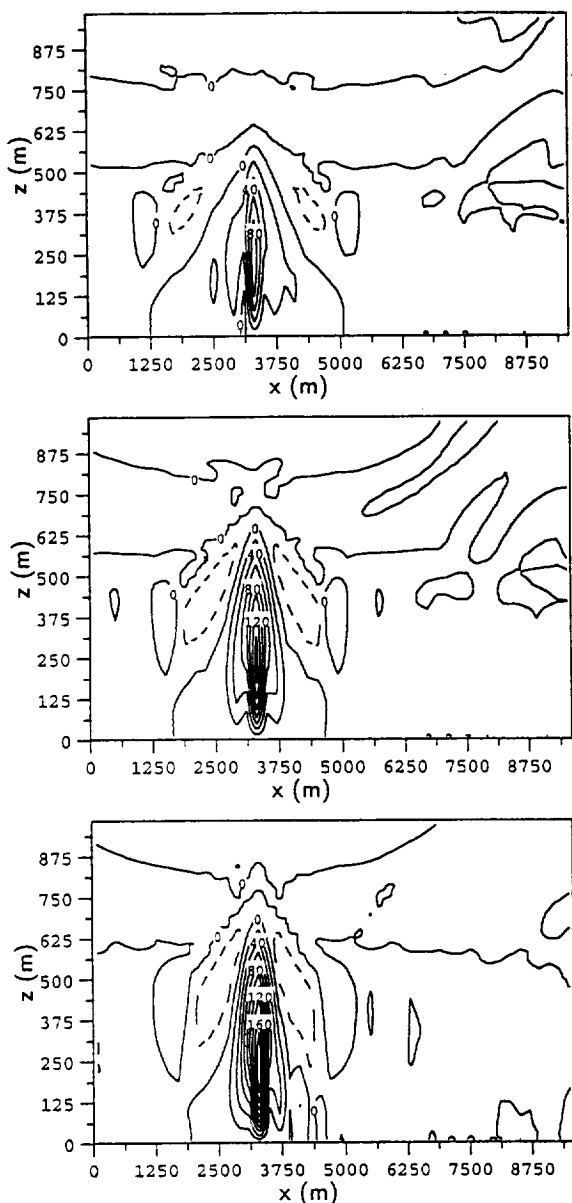


Figure 7: Mean vertical velocity (cm s^{-1}) for a 6400 m lead, 0° wind orientation: no microphysics and no radiation (top), includes microphysics (middle), includes microphysics and radiation (bottom).

Based on the present CRM results, it appears that the addition of microphysical and radiative processes has a greater impact upon the more energetic circulations, such as those associated with wider leads. Fig. 7 displays the results obtained for a 6400 m lead under three configurations. The first is with no additional physics; the second is with the addition of microphysical processes; and the third is with the addition of microphysical and radiative processes. Although the structure of the resulting plumes is not appreciably al-

tered, the depth and maximum vertical velocities do increase. For the first simulation, the plume reaches a depth of about 630 m. The addition of microphysics increases the plume depth by 60 m, and the further addition of radiative processes yields a plume height of 780 m. Thus, the inclusion of the additional physics increases the plume depth by nearly 25%.

The increase in circulation strength is due mainly to two factors. The surface latent heat flux is nearly one third the magnitude of the sensible heat flux. The amount of latent heat that is actually released through condensation depends upon the strength of the circulation and the initial moisture profile. This explains why microphysical processes typically have a greater impact upon larger leads, as their stronger circulations are able to lift the moist air higher, resulting in increased condensation and latent heating. For the simulations described herein, the initial moisture profile is set to 85% RH (with respect to ice). The addition of radiative processes results in a significant warming (through IR flux divergence) of the air directly above the lead. The radiation is also responsible for substantial IR cooling at the top of the plume, but the impact is more limited than caused by the heating, resulting in a net increase of the circulation strength.

ACKNOWLEDGMENTS

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The entrainment rate at the top of the atmospheric boundary layer is affected by many processes, such as radiation, evaporation, and boundary layer circulations. An intercomparison of radiatively driven entrainment and turbulence in a smoke cloud (Bretherton et al. 1997; hereafter B97) investigated an idealized case in which a convective boundary layer filled with radiatively active "smoke" is simulated by several 1D, 2D, and 3D models. Cloud-top cooling rates were chosen to be comparable to those observed in marine stratocumulus, while avoiding evaporative feedbacks on entrainment and turbulence that are also important in liquid water clouds. The radiative cooling rate had a specified dependence on the smoke profile, so that differences between simulations could only be a result of different numerical representations of fluid motion and subgrid-scale turbulence.

We used this case to test the University of Utah 1D turbulent closure model (UU TCM). The UU TCM also participated in a similar, but more complicated, intercomparison of simulations of a stratocumulus-topped boundary layer that motivated the smoke cloud intercomparison (Moeng et al. 1996). The UU TCM is the 1D version of the UU Cloud Resolving Model (Krueger 1988). The UU TCM is based on third-moment turbulence closure, which includes prognostic equations for first moments (including the horizontally averaged potential temperature and smoke mixing ratio), second moments (including the horizontal and vertical velocity variances and the vertical fluxes of sensible heat and smoke), and third moments (including the vertical fluxes of the horizontal and vertical velocity variances).

The initial boundary layer was assumed to be at constant potential temperature and uniformly filled with smoke (with a mixing ratio of 1) over its whole 700 m depth. Above the boundary layer the smoke concentration was set to zero. The standard requirement was that models should use a vertical grid size of 25 m and be run to simulate a period of 3 h.

Fig. 1 shows the 2-3 h average turbulent kinetic energy (TKE) budget for the UKMO 3D high-resolution large-eddy simulation (LES) model and the UU TCM.

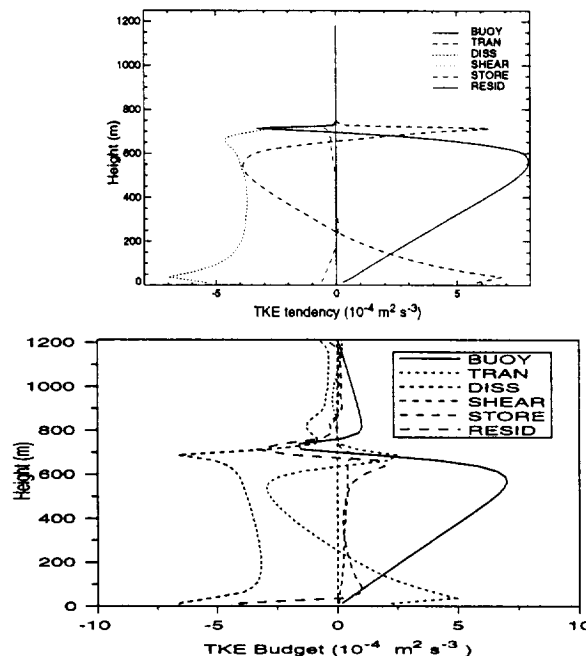


Figure 1: 2-3 h average TKE budget for the UKMO 3D high-resolution model (top; from B97) and the UU TCM (bottom).

The TKE budget for the UU TCM agrees well with that of the LES. The most notable differences occur above the boundary layer, where the UU TCM produces a spurious turbulent layer.

Figs. 2 and 3 show the potential temperature and smoke mixing ratio profiles averaged over 2-3 h for the UU TCM, while Figs. 4 and 5 show the corresponding turbulent flux profiles. When these are compared with the corresponding plots in B97, it is evident that the profiles of potential temperature, smoke, heat flux, and smoke flux from the UU TCM agree well with 3D large-eddy simulations involved in the intercomparison. The only significant differences occur above the boundary layer, where the UU TCM produces spurious fluxes of both heat and smoke.

In B97, results from numerous modeling groups around the world were compared with each other and

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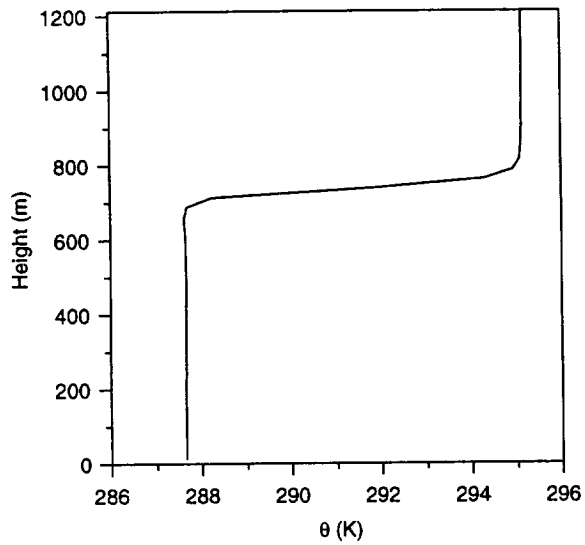


Figure 2: Potential temperature profile averaged over 2-3 h for the UU TCM.

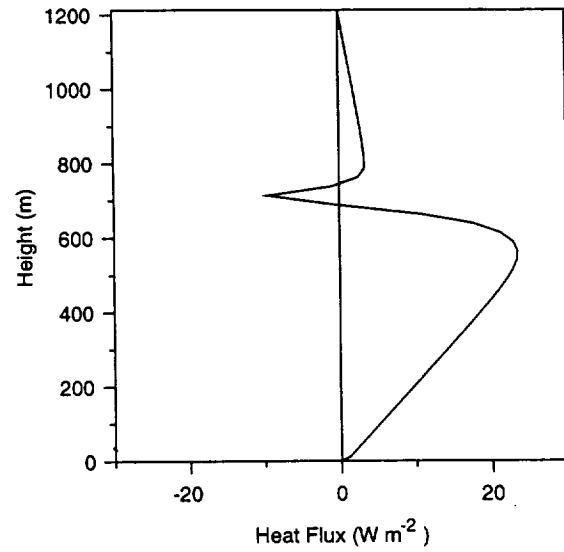


Figure 4: Heat flux profile averaged over 2-3 h for the UU TCM.

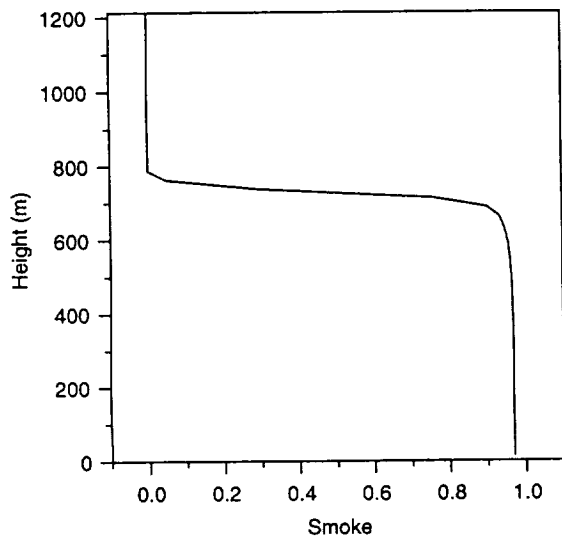


Figure 3: Smoke mixing ratio profile averaged over 2-3 h for the UU TCM.

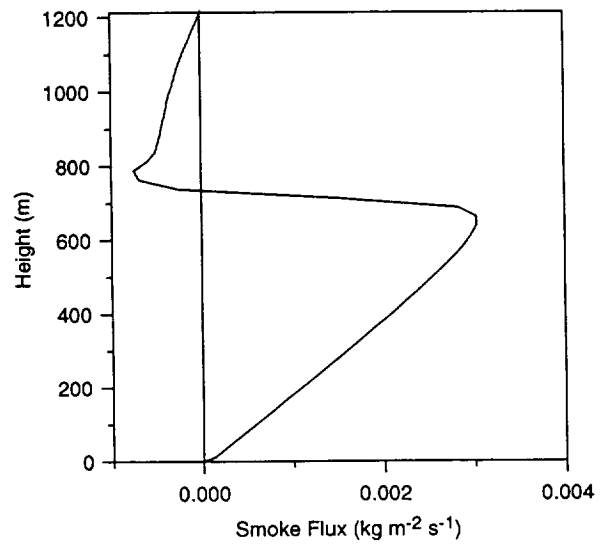


Figure 5: Smoke flux profile averaged over 2-3 h for the UU TCM.

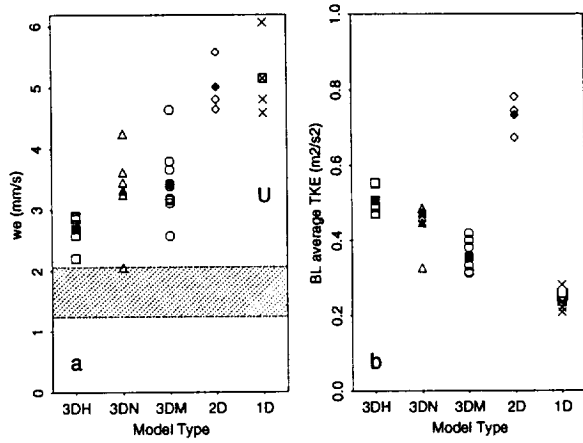


Figure 6: Scatterplot of 2-3 h averages of (a) w_e , (b) TKE_{blav} , partitioned by group. Filled symbols (crossed square for 1D) indicate means for each group. "U" shows the result for the UU TCM. Hatched zone in (a) indicates prediction based on laboratory analog. (After B97.)

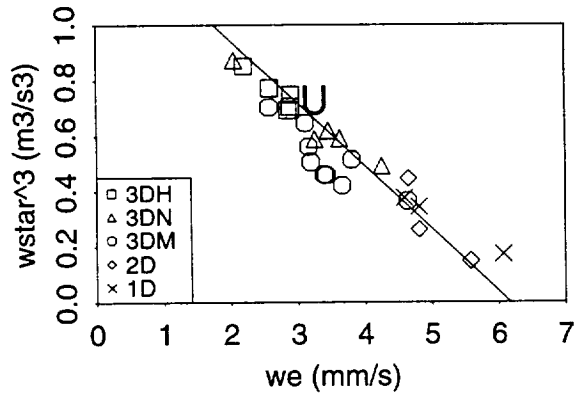


Figure 7: Scatterplot of w_{star}^3 vs. w_e for all models. "U" shows the result for the UU TCM. The line is the prediction of (11) in B97. (After B97.)

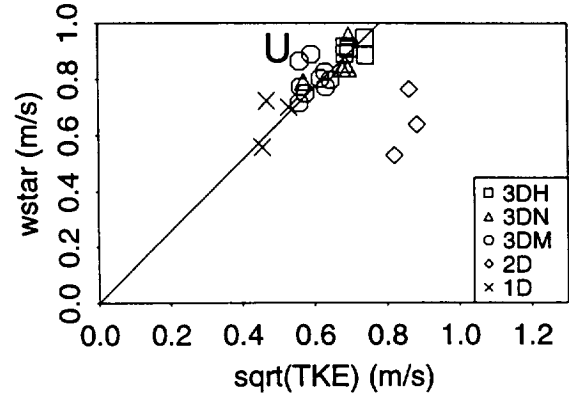


Figure 8: Scatterplot of w_{star} vs. $\sqrt{TKE_{blav}}$ for all models. "U" shows the result for the UU TCM. Best-fit line is derived from the 3DH models. (After B97.)

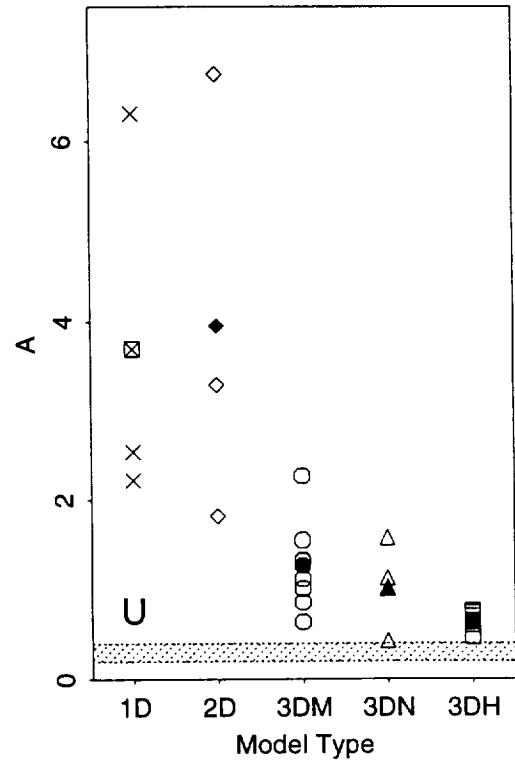


Figure 9: As in Fig. 6a, except for A . (After B97.)

with a previously investigated laboratory analog to the smoke cloud. We will compare the results from the UU TCM to these results.

Fig. 6 is a scatterplot of 2-3 h averages of the entrainment velocity, w_e , and the TKE averaged over the boundary layer, TKE_{blav} , partitioned by model group: three-dimensional, high resolution (3DH); three-dimensional, standard resolution with non-monotone advection (3DN) or monotone advection (3DM); two-dimensional (2D) and one-dimensional (1D). The symbol "U" shows the result for the UU TCM. The hatched zone in (a) indicates the prediction based on the laboratory analog (see B97 for details). The entrainment velocity for the UU TCM is much lower than that for other 1D models, and is in the same range as that of the 3D standard-resolution models (3DN and 3DM). Whether this is a robust feature of third-moment turbulence closure models, or is an artifact of this particular model, remains to be determined.

Fig. 7 is a scatterplot of w_*^3 vs. w_e for all models, where w_* is the convective velocity scale. The line is the prediction of Eq. (11) in B97, which was derived under the assumption that the boundary layer is well-mixed. The point for the UU TCM falls near the theoretical line along with those from the 3D models.

Fig. 8 is a scatterplot of w_* vs. $\sqrt{\text{TKE}_{\text{blav}}}$ for all models. The best-fit line is derived from the 3DH models. The point for the UU TCM is above the best-fit line. The reason for this is not understood.

Fig. 9 is a scatterplot of 2-3 h averages of A , the entrainment efficiency (proportional to w_e/w_*^3), partitioned by model group. The hatched zone shows the range suggested by the laboratory experiments. (See B97 for details.) The value for the UU TCM is similar to that of the 3D models.

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Despite the recognition that Arctic stratus clouds (ASC) have a significant effect on the surface energy budget of the Arctic, and thereby on the sea ice thickness, the formation mechanism(s) of these clouds are still uncertain. This is largely due to the obvious logistical difficulties of making extensive observations in the Arctic. However, limited observations have been made, including during the Arctic Stratus Experiment in the Beaufort Sea during June 1980. These and other observations of the structure of the summertime arctic boundary layer were summarized by Curry et al. (1988). They noted that ASC differ in several aspects from marine subtropical stratus clouds. ASC often occur in multiple layers. The lowest layer may rest on the surface. The upper cloud layer or layers are decoupled from the stable surface layer, thus their source of water vapor does not appear to be the surface.

The formation of ASC was studied by Curry and Herman (1985) using large-scale budgets derived from operational analyses. They concluded that ASC form as relatively warm and moist air flows into the Arctic from lower latitudes and is cooled radiatively and by contact with the colder surface of the sea ice and ocean.

Three mechanisms have been proposed to explain the layering of ASC. The 1D turbulence closure-radiative transfer model results of Herman and Goody (1976) support the idea that solar absorption warms the interior of a surface-based cloud and thereby forms a clear layer between two cloudy layers. Tsay and Jayaweera (1984) proposed that the upper cloud layer is formed by very weak ascent or entrainment, while the surface cloud layer is an advective fog. McInnes and Curry (1995) suggested that the upper cloud layer is maintained by cloud-top radiative cooling, while the lower cloud layer is formed by radiative cooling at the base of the upper mixed layer.

Modeling studies that used 1D second-moment turbulence models have reproduced many aspects of the structure of the summertime Arctic boundary layer (Finger and Wendling 1990; McInnes and Curry 1995).

In this preprint, we present some preliminary results on the sensitivity of the formation and structure of simulated Arctic summertime boundary-layer clouds to large-scale vertical velocity and drizzle using the University of Utah 1D turbulent closure model (UU TCM). The UU TCM includes third-moment turbulence closure, a drizzle parameterization, and an interactive radiative transfer scheme. The UU TCM is the 1D version of the UU Cloud Resolving Model (Krueger 1988).

The UU TCM has been tested against 3D large-eddy simulation models in two recent model intercomparisons that involved simulations of cloud-topped boundary layers (Moeng et al. 1996; Krueger et al. 1999). In both intercomparisons, the UU TCM's boundary layer vertical structure and entrainment rate were nearly the same as those obtained from the large-eddy simulation models.

Here we present the results of two sets of integrations. In one set (Cases B, W1, W2, and W3) only the large-scale vertical velocity \bar{w} is varied, while in the second set (Cases Cases B, D1, and D2), only the drizzle parameterization is changed. Case B is the baseline case against which the others are compared. Table 1 lists the cases.

The drizzle parameterization labeled "Chen" in Table 1 follows Chen (1996) and Wang and Wang (1994). The parameters we used lead to an autoconversion threshold of 0.38 g kg^{-1} . We chose to use an autoconversion rate of 0.0002 s^{-1} which corresponds to Wang and Wang's "light drizzle" case. The parameterization labeled "Kessler" uses Kessler's warm rain microphysics as described in Krueger (1988) except that the autoconversion threshold is reduced to 0.3 g kg^{-1} . The autoconversion rate remains unchanged at 0.001 s^{-1} .

The initial conditions for our integrations are based on the observations of an Arctic summertime boundary layer obtained on June 28, 1980, during the Arctic Stratus Experiment (McInnes and Curry 1995). In the integrations, the effective radius of the cloud droplets is 7 microns, the surface albedo is 0.55, the solar zenith angle is 74 degrees, and the vertical grid size is 25 m. Each case was run for 10 hours of physical time.

Fig. 1 presents time-height plots of cloud water mix-

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Table 1: List of cases.

Case	\bar{w} (mm s ⁻¹) at $z = 1000$ m	Drizzle Parameterization
B	0.0	Chen
W1	-0.2	Chen
W2	-0.5	Chen
W3	-4.0	Chen
D1	0.0	Kessler
D2	0.0	(none)

ing ratio for Cases B, W1, W2, and W3. The baseline case, under conditions of no large-scale vertical motion, contains two layers of clouds: a stable fog layer near the surface and a stratus cloud in an elevated mixed layer. The top of the upper cloud layer ascends slowly due to entrainment. Cases W1, W2, and W3 show that large-scale subsidence delays the formation of the upper cloud layer, and increases the liquid water content of the fog layer.

Fig. 2 includes time-height plots of cloud water mixing ratio for Cases D1 and D2. These should be compared to the baseline case plot in Fig. 1. The main impact of changing the drizzle parameterization from Chen to Kessler is to slightly reduce the peak cloud water mixing ratios in the upper cloud layer, while neglecting drizzle slightly increases them. The impact of changing or not including the drizzle parameterization is considerably greater in the fog layer due its larger peak cloud water mixing ratios. It should be noted that in the cases with drizzle, the downward turbulent flux of water at the surface is comparable to the drizzle flux.

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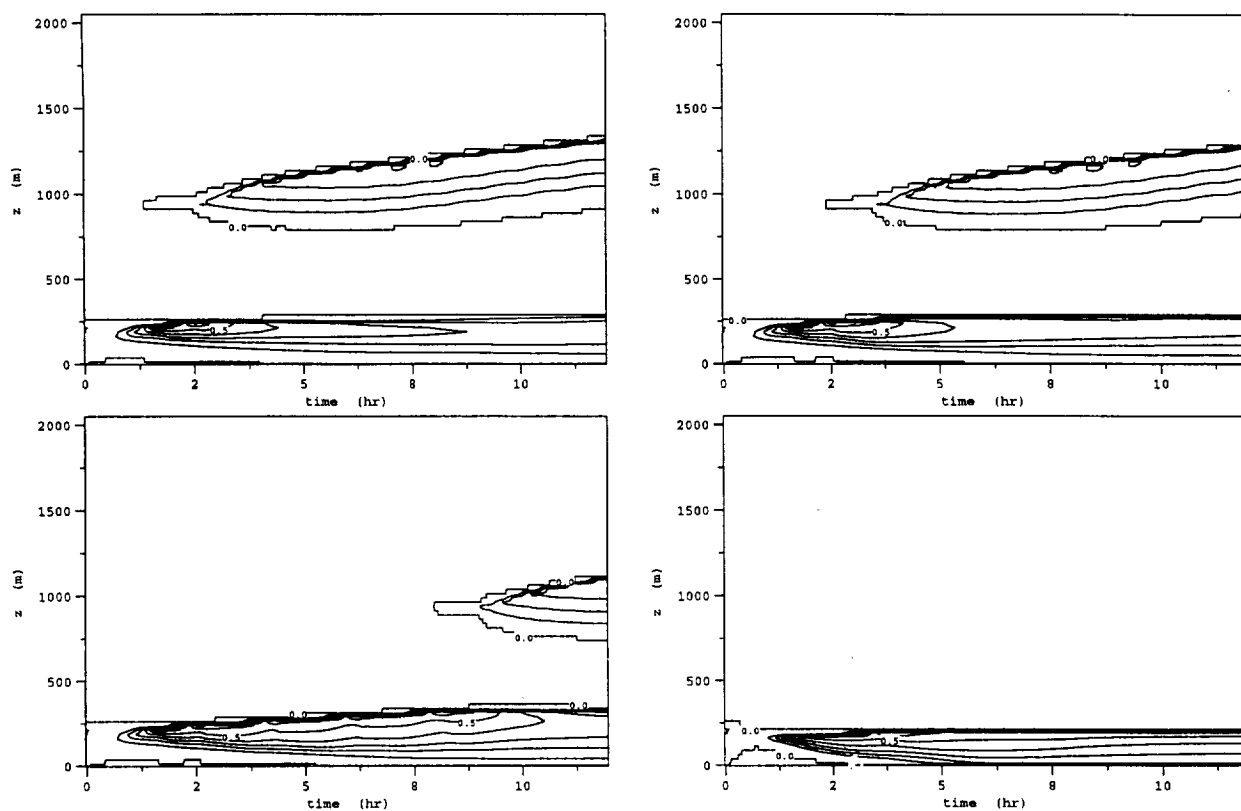


Figure 1: Time-height plots of cloud water mixing ratio (g kg^{-1}) for Cases B (upper left), W1 (upper right), W2 (lower left), and W3 (lower right).

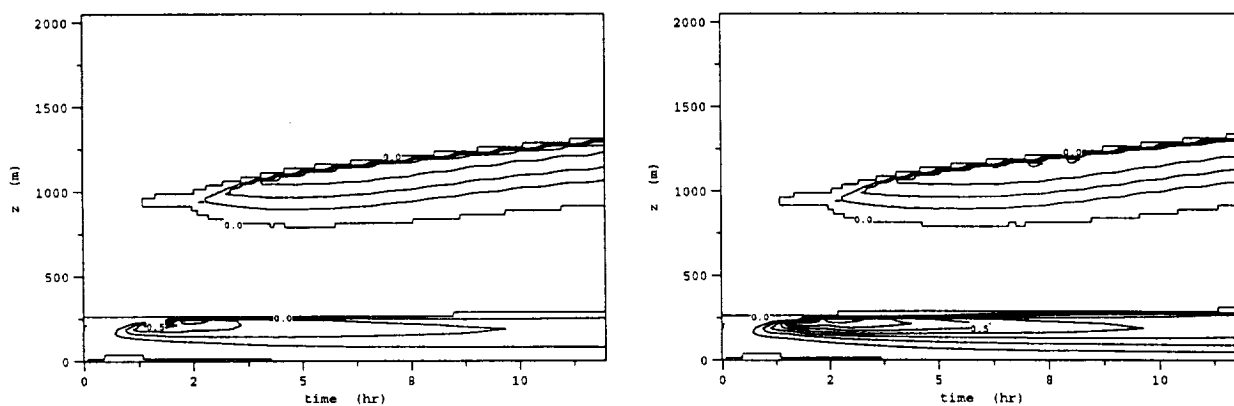


Figure 2: Time-height plots of cloud water mixing ratio (g kg^{-1}) for Cases D1 (left) and D2 (right).